

A MARTIAN GLOBAL GROUNDWATER MODEL

Alan D. Howard, Department of Environmental Sciences, University of Virginia, Charlottesville, VA 22903

A global groundwater flow model has been constructed for Mars to investigate hydrologic response under a variety of scenarios, improving and extending earlier simple cross-sectional models [1]. The model is capable of treating both steady-state and transient flow as well as permeability that is anisotropic in the horizontal dimensions. A single near-surface confining layer may be included (representing in these simulations a coherent permafrost layer). Furthermore, in unconfined flow locations of complete saturation and seepage are determined. The flow model assumes that groundwater gradients are sufficiently low that Dupuit conditions are satisfied and the flow component perpendicular to the ground surface is negligible. In spherical coordinates the governing equation is:

$$\frac{\partial}{\partial \lambda} \left\{ h_e \left[\frac{K_{\lambda}}{\cos \Phi} \frac{\partial h}{\partial \lambda} + K_{\lambda \Phi} \frac{\partial h}{\partial \Phi} \right] \right\} + \frac{\partial}{\partial \Phi} \left\{ h_e \cos \Phi \left[\frac{K_{\lambda \Phi}}{\cos \Phi} \frac{\partial h}{\partial \lambda} + K_{\Phi} \frac{\partial h}{\partial \Phi} \right] \right\} \\ = R^2 \cos \Phi \left[S \frac{\partial h}{\partial t} - Q \right],$$

where λ is longitude, Φ is latitude, h is the local hydraulic head, R is the planetary radius (assumed constant), S is the specific storativity (confined portions of the aquifer) or specific yield (unconfined areas), Q is distributed vertical recharge, K_{λ} , K_{Φ} , and $K_{\lambda \Phi}$ are the three components of the symmetric hydraulic conductivity tensor, and h_e is an *effective* aquifer thickness which, in the case of an aquifer with K that does not change vertically (termed the *uniform* case) is equal to $(h_u - h_l)$, where h_u and h_l are the elevations of the top and base of the aquifer, respectively. In confined situations h_e equals $(h_c - h_l)$, where h_c is the base of the confining layer. In some simulations K was assumed to decrease exponentially with depth from the surface at elevation h_s (the *exponential* case), giving an effective thickness of

$$h_e = \frac{1}{\alpha} e^{-\alpha(h_s - h_u)}$$

where α is a decay constant (in these simulations the aquifer is assumed to be unbounded at depth). In these simulations storativity was assumed also to decrease exponentially with depth with the same decay constant. In unconfined regions h_u equals the water table elevation, but in confined circumstances h_u is the base of the confining layer, h_c .

Surface elevations were taken from the new 1:15M Mars topographic map. Two sets of elevations were recorded for each grid-point location, the first (the *nominal* set) being the local elevation, which, however, was adjusted upwards to the level of the surrounding uplands if the gridpoint occurred within a crater or in an erosional channel. However, locations in craters large enough to encompass more than one grid points (e.g., Hellas and Argyre) were represented by the local elevation. The second set of elevations (the *fluvial* set) was taken in the same manner as the nominal set except that the elevation was adjusted downwards to the elevation of the bottom of any nearby erosional channel. Thus the nominal set is assumed to represent the pre-channeling landscape and the fluvial set to represent the present fluvial base level. Examination of both outflow and the larger cratered-terrain valley networks reveals that most flowed down essentially the present regional topographic gradients. The few local disparities between inferred channel flow directions and topographic maps (e.g. Nirgal, Dueteronilus, Mangala) are probably as likely to be due to uncertainties in topographic mapping as to regional tectonism. The major exception is the Tharsis volcanic construct, largely postdating the development of valley networks. However, flow directions of most outflow channels associated with the margins of Tharsis appear to be consistent with present topographic gradients, suggesting that most of the marginal updoming and normal faulting in Valles Marinaris predated the outflow channels. In addition, the direction and degree of development of

the valley networks suggests that the north-south topographic discontinuity (highlands-lowlands transition) had essentially its present configuration prior to the development of the late Noachian valley networks. Thus the present topography reasonably can be considered to be indicative of conditions during channel development.

In simulations with a finite aquifer thickness, the local base of the aquifer, h_1 , was set to a given elevation below the local fluvial elevations. In the present simulations the aquifer thickness below the fluvial elevations was assumed to be areally uniform.

The flow equations were solved using a finite difference method employing 10-degree spacing of latitude and longitude. A successive over-relaxation method (SOR) was used for steady-state solutions and a Crank-Nicholson method with iterative solution at each time step was used for transient simulations. The correctness of the solution method was checked in part by confirming mass conservation. In addition, an independent, finite element flow model was also constructed for steady-state conditions and compared to the finite difference model. The finite element model represented the planet by triangular planar elements matched at their edges and the flow equations were cast in cartesian form. The two models gave essentially identical solutions.

The initial guess for steady state flow simulations was a level water table with an elevation equal to the lowest surface elevation on the planet (the bottom of Hellas). If the base of the aquifer was above this level then the water table was set to the base of the aquifer. Steady state simulations assumed a constant recharge of the aquifer from the surface. In most cases an areally uniform recharge rate was assumed, but some simulations were conducted that assumed recharge to be a function of latitude and/or elevation:

$$Q = Q_0 [\cos\Phi] \{1 + \beta h_s\}$$

where Q_0 is a nominal recharge rate, and β is an input parameter. The bracketed latitude correction is optional. Recharge to fixed-head and confined portions of the aquifer was set to zero. The only location on the planet that was specified as fixed-head at the beginning of the iterative solution was the bottom of Hellas. Locations of saturation and, therefore, seepage were determined during the SOR iterative solution by identifying during each iteration locations where the predicted head, h , was greater than the surface elevation, h_s . Such locations were converted to fixed-head for succeeding iterations until the steady-state solution was achieved. A similar procedure was used to identify confined portions of the aquifer wherever the predicted head was greater than the bottom of the confining layer, h_c .

Transient flow simulations started from steady state conditions, and involved either of two scenarios: 1) *Draining*: The aquifer system is initially equilibrated with a specified recharge rate and then drained with no further recharge, and 2) *Filling*: The aquifer system initially has no available water (i.e., it is set to the initial conditions for the steady state iterations as described above) and the aquifer then fills towards steady state with a specified constant recharge.

Steady-state flow simulations. Steady-state flow simulations have been conducted to investigate effects of parameter variation, particularly the ratio of recharge rate, Q , to intrinsic permeability, k . Hydraulic conductivity has been assumed to be isotropic and uniform in these simulations. Another set of runs examines the effects of a confining permafrost layer extending from the poles to a variety of latitudes. Finally some simulations have been conducted to examine the effects of spatially variable recharge and anisotropic permeability.

Transient flow simulations. A two-dimensional flow model had previously been used to calculate the length of time for martian aquifers to drain following cessation of recharge [1]. These figures are updated here using the global flow model. Results from a number of simulations run with different sets of model parameters were analyzed by multiple regression to determine estimating equations for the length of time for specified percentages of filling, t_f , or draining, t_d , of the aquifer. The estimating equation for draining flows is given by

$$t_d = C_d \frac{v\eta}{kg} G^{\delta} V_r^{\gamma}$$

where C_d is a coefficient (see table below), ν is the water viscosity, η is the porosity, k is the intrinsic permeability (note that hydraulic conductivity, K equals kg/ν), g is the gravitational constant, G is the average hydraulic gradient at the start of aquifer draining, and V_r is the ratio of the available water volume (water above the lowest discharge point in Hellas) to the total water volume (including water below the lowest discharge point) at the start of aquifer draining. The exponents take the values of $\delta=-1.3$ and $\gamma=1$ for the uniform case and $\delta=-1.0$ and $\gamma=-1.4$ for the exponential case. Similarly, for the case of filling flows:

$$t_f = C_f \eta Q \left[\frac{\nu}{kg} \right]^\epsilon D^\gamma$$

where D is the depth from the surface to half-value of the permeability ($0.693/\alpha$) and the exponents take the values of $\sigma=-0.8$, $\epsilon=0.2$ and $\gamma=0$ for the uniform case and $\sigma=-0.8$, $\epsilon=0.2$ and $\gamma=1.3$ for the exponential case. The times to filling or draining can be specified in two ways: 1) the time required to drain or fill a specified percentage of the total available water volume (constants C_{dv} and C_{fv}); or 2) the time required for the total discharge of water to decrease or increase by a specified percentage (constants C_{dq} and C_{fq}). The values of the coefficients are given below:

Percent Change	Uniform Case				Exponential Case			
	C_{dv}	C_{fv}	C_{dq}	C_{fq}	C_{dv}	C_{fv}	C_{dq}	C_{fq}
10	13.59	0.545	4.33	1.94	69	0.33	29.4	8.49
25	36.9	1.45	10.3	3.58	255	0.84	90.7	1.34
50	146	3.07	31.6	5.61	1148	1.78	306.3	2.67
75	463	5.32	108.4	7.91	5115	2.96	1195	3.92
90	904	7.73	298	9.91	30019	4.17	3346	5.11

In using this table Q is expressed in cm/yr, D in km, k in darcies, g in cm/sec², ν in cm²/sec, and times, t , in 10⁶ yr.

The times for aquifer draining are generally comparable to those reported for the two-dimensional simulations [1]. For example, for the uniform permeability case with $\eta=0.2$, $k=1$, $G=0.0017$ and $V_r=0.67$ the 2-D simulations with a representative aquifer length of 3000 km indicate the time to 75% draining as 5.6×10^6 yr, and the present simulations indicate 12×10^6 yr for the uniform aquifer and 48×10^6 yr for the exponential aquifer. However, the draining times increase much more strongly with percent draining than in the case of the 2-D simulations because the length-scale increases as draining progresses and the water must flow towards more distant exit points.

Times for aquifer recharge are essentially inversely proportional to recharge rate. Furthermore, recharge generally occurs much more rapidly than draining for comparable regolith permeability. For example, recharge at a rate of 1 cm/yr will fill the regolith to 75% of its steady-state capacity in only 150,000 years (assuming $\eta=0.2$ and $k=1$) compared to tens of millions of years for a comparable percentage of draining. In addition, recharge rates are only weakly dependent upon aquifer permeability.

References:

- [1] Howard, 1990, *Rpts. Planetary Geol. Geophys. Prog.*, NASA TM 4210, 342-344.